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Chapter 3 South Asian Summer Monsoon Variability in a Changing Climate

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The Monsoon and Climate Change

Disclaimer

Chapter 3

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Abstract

This chapter provides a succinct review on the current understanding of the South Asian summer monsoon and the ability of present-day climate models to represent its variability in a changing climate. Beginning with a processes-based review of the large and regional scale aspects of the monsoon precipitation climatology, the systematic model errors in precipitation and monsoonal diabatic heating are also highlighted. In climate models, certain necessary conditions for representation of synoptic systems, boreal summer intraseasonal variability, and the ENSO-monsoon teleconnection are presented. This is followed by discussion of the improved (or lack thereof) performance of climate models in simulating natural modes of variability. Lastly, we evaluate the ability of models to simulate the observed long-term declining trend in the seasonal mean monsoon rainfall, including possible mechanisms for this trend. Despite dedicated efforts, there is a lack of substantial improvement in monsoon modeling, which in our view is due to the lack of high-quality observations (atmosphere and ocean) over the monsoon-influenced regions to constrain the model physics. *Our conclusion is that without such an observational effort, improving the physical processes in numerical models will be severely limited.*

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1. Introduction

Over the past six decades, sustained research from observations, models, and theory have led to a consensus that monsoons arise due to complex interactions among ocean, atmosphere and land components of the climate system. The annual variation in solar radiative forcing provides the necessary thermodynamic conditions for the development of regional monsoons. For instance, in the annual cycle, the Asian-Australian monsoon (AAM) system can be regarded as the seasonal displacement of the large-scale Inter Tropical Convergence Zone (ITCZ) that is anchored by the north-south migration of the Indo-Pacific warm pool (regions where sea surface temperature is > 28^{o}C). Despite measurable progress in our overall understanding of the monsoons (e.g., Webster et al. 1998), and exponential growth in "computing facilities", modeling the seasonal displacement of the warm pool (Annamalai et al. 2014) and the associated rainfall patterns have met with limited progress (Sperber et al. 2013). Systematic model errors in the climatological basic states cascade into the simulation of subseasonal (Sperber and Annamalai 2008), interannual (Turner et al. 2005; Annamalai et al. 2007) and long-term variations of the monsoons (e.g., Annamalai et al. 2013). Additionally, it is recognized that the monsoon phenomenon is intrinsically complex, and that any error in one component of the simulated climate system cascades into the other components. Thus, future projections of mean monsoon and its spectrum of variability by the state-of-the-art climate models exhibit large uncertainties (Turner and Annamalai 2012).

Compared to six decades ago, present-day climate scientists have access to observations from multiple platforms (in-situ, satellite, reanalysis, dedicated field experiments, etc.) and powerful computers for analyzing this data. This has led to innumerable scientific publications pertaining to various aspects of the AAM. Yet, the lack of significant progress in monsoon modeling raises questions: (i) Is our understanding adequate enough to model the monsoons? (ii) Is the current level of observations adequate enough to constrain the models? (iii) Have we reached the upper-limit of understanding and modeling the monsoons? The present chapter provides a succinct review of the current understanding of the South Asian monsoon and the ability of present-day climate models in representing its variability in a changing climate.

We begin with a brief review of the processes that may be important for the development and maintenance of the mean monsoon precipitation (Section 2). In Section 3) we discuss the subseasonal variations (synoptic and intraseasonal variations), while Section 4 is devoted to interannual variations with a particular emphasis on the ENSO-monsoon association. The long-term declining tendency in regional monsoon rainfall is discussed in Section 5. In each section, after reviewing observational aspects, we will present and discuss their representation in climate models, and close with cautionary notes on both observational and climate model uncertainties. Finally, in Section 6 we provide our perspective for future directions in monsoon research.

2. Mean monsoon

2.1 Large-scale aspects: In the annual cycle, based on thermodynamical principles, intense solar heating during boreal spring and early summer requires deep convection poleward of the Equator. Figure 1a shows the satellite-based observed precipitation and SST climatology during boreal summer. Figure 1b shows sea level pressure (shaded) and 850 hPa wind climatologies constructed from ERA-interim reanalysis. The intense solar heating in late spring and early summer anchors the north-northwest migration of the warm pool. This, in conjunction with land-surface heating, lead to the formation of a diagonally oriented low-pressure area that extends from Arabian peninsula to the tropical western Pacific (TWP; Fig. 1b). Figure 1a shows that ocean points in the region 10^{O}S-20^{O}N; 70^{O}-160^{O}E experience high-mean SST (> 28^{O}C) and intense rainfall (>9 *mm/day*), supporting the thermodynamic view that the tropical SST distribution determines the low-level moist static energy (MSE) that in turn anchors moist convection (Neelin and Held 1987; Raymond 1995). At the low- level, the Mascarene High in the Southern Hemisphere and the low-pressure region in the Northern Hemisphere, termed the monsoon trough, are connected by the cross-equatorial flow (Fig. 1b) that feeds moisture to monsoon convection. More discussions on the large-scale features of mean monsoon can be found in Turner and Annamalai (2012) and Sperber et al. (2013).

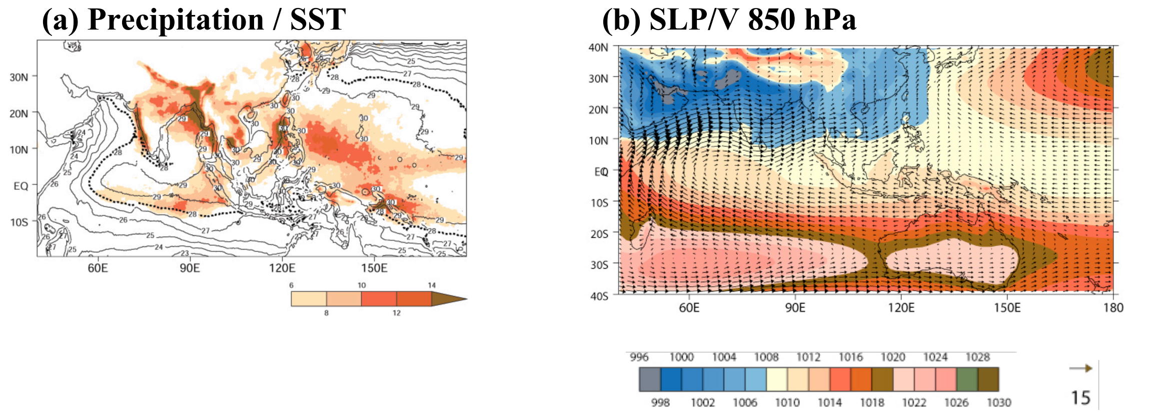


Figure 1: Seasonal mean (JJAS) climatology of: (a) precipitation (mm/day; shaded) and SST (^{O}C, contour), and (b): 850hPa wind (m/s) and sea level pressure (hPa; shaded). Unit vector is also shown. Precipitation is taken from TRMM and SST is from TRMM Microwave Imager (TMI) for the period 1998-2012; SLP and wind are taken from ERA-Interim for the period 1989-2012.

2.2 Regional-scale aspects: At regional scales, the poleward migration of the warm pool (28^{o}C isotherm reaches as far as 28^{o}N; Fig. 1a) has specific dynamical and thermodynamical implications for the monsoon over the TWP that include: (i) the northeastward extension of the subtropical high (Fig. 1b; Wu and Wang 2001); (ii) the eastward extension of the monsoon trough and low-level westerlies (Fig. 1b); and (iii) enhanced surface fluxes and boundary layer entropy (Raymond 1995). Despite the fact that SST-convection relationship is nonlinear and complex over the warm pool region (Lau et al. 1997), a necessary condition is that coupled models need to realistically simulate the zonal and meridional extent of the warm pool over the TWP to anchor in-situ convection (Annamalai et al. 2014).

Over South Asia, convection–Rossby-wave interactions (Rodwell and Hoskins 1996), in conjunction with warmer SST over the Bay of Bengal and cooler SST over the Arabian Sea (Shenoi et al. 2000), help set up an east/west asymmetry of precipitation (Fig. 1a). Prior to monsoon development, the northern Indian Ocean is the warmest of all tropical oceans (Schott and McCreary 2001). During the monsoon season, the SST drops to about 23-24^{o}C along the Somali coast (Fig. 1a), primarily due to the upwelling of cold water by the cross-equatorial low-level jet (Fig. 1b). Subsequent horizontal advection by ocean currents, in conjunction with evaporative cooling, results in colder SST over the central

Arabian Sea (McCreary et al. 1993). In the absence of such ocean-atmosphere processes, SST over the Bay of Bengal remains very high. Given the theoretical evidence that the intensity of the low-level jet along the Somalia depends on an accurate representation of the East African Highlands (Hoskins and Rodwell 1995), horizontal and vertical resolutions employed in the atmospheric model component will have an impact on the intensity of upwelling and SST cooling off Somalia. Thus, we recognize that the cross-equatorial jet and monsoon convection are intrinsically tied to each other. From a vorticity perspective, regions south of the axis of the low-level jet experience anticyclonic flow (Fig. 1b) and descent (not shown). As a consequence, the equatorial western Indian Ocean (0-10^{o}N; 60^{o}-70^{o}E) experiences minimum rainfall despite the presence of high-mean SST, implying dynamical rather than thermodynamical control.

Both high SST and steep orography along the ocean-land boundaries play a critical role in the rainfall distribution (Fig. 1a). These determine the locations of maximum rainfall in the vicinity of Western Ghats, the northern bay adjoining the Arakan Range in Burma, and the South China Sea adjoining the northern mountain chain of the Philippines. Note that orographic only forced rainfall along the foothills of the Himalayas and Tibet is modest. The concentrated rainfall (> 18 *mm/day*) over the central-northern Bay of Bengal may be attributed to dynamics (monsoon trough), thermodynamics (high mean SST and associated MSE) and orographic forcings (Annamalai et al. 2014).

2.3 Mean monsoon in CMIP3/5 models: In climate models, the mean state of the monsoon, particularly the large-scale circulation features are better represented than regional-scale precipitation (Sperber et al. 2013). For instance, the Asian summer monsoon is comprised of multiple rainfall maximum zones (Fig. 1a) that represent (i) the Indian monsoon (70^{o}E-100^{o}E, 10^{o}N-25^{o}N), (ii) the tropical western Pacific (110^{o}E-150^{o}E, 10^{o}N-20^{o}N) and (iii) the eastern equatorial Indian Ocean (10^{o}S-0, 90^{o}E-110^{o}E). Because these centers do not respond in unison to any internal or external forcing (Annamalai and Sperber 2005; Annamalai et al. 2007; Annamalai 2010), and influence each other at all time scales, a realistic representation of these regional centers is a pre-requisite if the models are to adequately capture the monsoon variability in a changing climate.

As seen in Fig. 2 for summertime precipitation, the multi-model mean (MMM) error relative to GPCP observations has shown little improvement in CMIP5 as compared to CMIP3 (Sperber et al. 2013). For the Asian summer monsoon the MMM monsoon rainfall is underestimated over South Asia and the central-eastern equatorial Indian Ocean, and overestimated over the western equatorial Indian Ocean and tropical West Pacific. The indication is that systematic model errors persist over the regional rainfall zones in subsequent generations of climate models. On a positive note, the increased model resolutions in CMIP5 improve the representation of orographic-induced rainfall (Sperber et al. 2013). For the austral summer monsoon the rainfall is excessive over most of the Maritime Continent, and deficient over northern Australia (not shown). One implication from our analysis is that uncertainties in future projections (e.g., IPCC 2013) of AAM mean rainfall may not have been reduced from CMIP3 to CMIP5 given the persistence of present-day mean state errors.

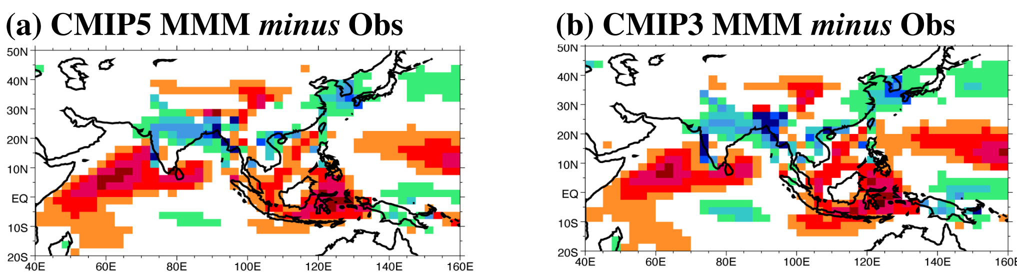
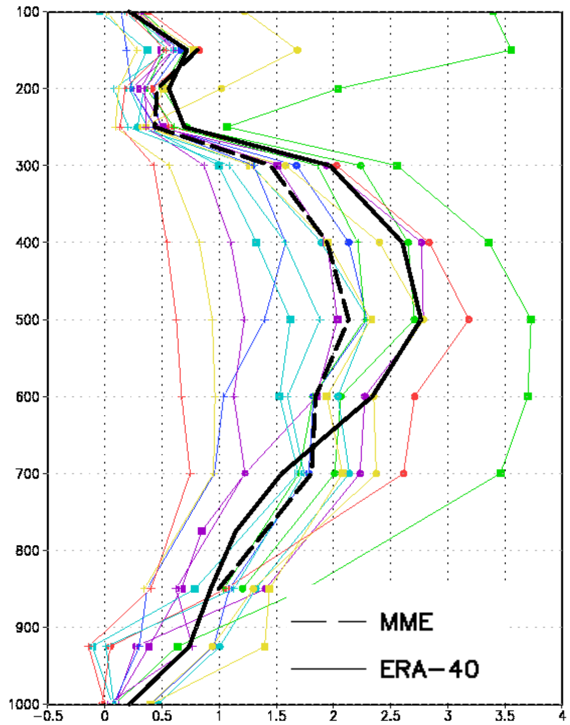


Figure 2: Boreal summer mean precipitation climatology difference (mm/day) between the present-day CMIP/55 multi- model-mean (MMM) and GPCP observations. After Fig. 1 of Sperber et al. 2013

The positive rainfall errors over the western Indian Ocean as well as over the Maritime Continent persist throughout the annual cycle (Annamalai 2014). Initial investigation has revealed the extent of the monsoon errors and their manifestation in the various model components. The errors in precipitation induce errors in wind-stress that subsequently impact ocean currents and thermocline depths (Nagura et al. 2013), and they impact the simulated SST (Annamalai et al. 2014b).

Another constraint to model improvement is "observational uncertainty". For example, an initial examination of the differences in boreal summer rainfall climatologies between two frequently used rainfall products (GPCP and CMAP) for model validation (not shown, see Fig. 1f of Sperber et al. 2013) indicates errors that are similar to the CMIP5/3 MMM precipitation differences seen in Figs. 2a and 2b.

As mentioned above, the regional distribution of mean monsoon precipitation is *likely* to arise due to complex interactions among dynamical, thermodynamical, and orographic forcings. Given that the interaction between large-scale motions and convection is inherent for the AAM system, a collective measure of all forcings that make up the monsoon rainfall can be inferred by the models' ability to represent the vertical profile of the diabatic heating. Figure 3 shows the vertical profile of diabatic heating (*Q*), area-averaged over South Asian monsoon region (5^{o}-25^{o}N, 60^{o}-100^{o}E) during boreal summer from CMIP5 models. *Q* is computed as a residual of the thermodynamic energy equation, following the approach used by Hoskins et al. (1989) and Nigam et al. (2000). The vertical structure of Q dictates the efficacy of the divergent circulation that either exports or imports MSE Among the models, large diversity exists both in terms of vertically integrated *Q* amplitude as well as its vertical structure that are expected to influence the three-dimensional circulation (Hoskins and Wang 2006). In the reanalysis, as expected over deep convective regions, *Q* peaks around mid-troposphere (500-400 hPa). Many models, on the other hand, tend to have maxima at the mid-troposphere but their simulated amplitude is overestimated in the lower troposphere (900-700 hPa) and underestimated in the layer 700-300 hPa, a feature readily apparent in the multi model mean composite (black long-dashed line; Cherchi et al. 2014). Some outliers, such as CSIRO-Mk3-6-0 and ACCESS1-3, do not show any appreciable vertical structure since the simulated monsoon over South Asia is virtually absent in those models (Sperber et al. 2013). TRMM observations indicate that over the monsoon region the stratiform rainfall contribution to the *Q* intensity is about 40% (Schumacher et al. 2004). In contrast, most of the CMIP3 models produce too much convective (95% of the total) and too little stratiform precipitation (Dai 2006). Given the persistence of systematic errors (Fig. 1), we speculate that errors in the partitioning of total rainfall into convective-stratiform may still persist in CMIP5. Furthermore, in CMIP5 models, higher *Q* intensity at lower tropospheric levels may be attributed to misrepresentations in shallow convection.



Diabatic heating (*Q*)

Figure 3: Vertical distribution of *Q* estimated from CMIP5 and ERA reanalysis (solid line). The multi-model-mean composite (dashed line) is also shown (from Cherchi et al. 2014)

To understand the models' ability to simulate the annual cycle of precipitation, cumulative and fractional accumulation methods are applied (Sperber and Annamalai 2014) using pentad precipitation. This approach provides an advantage over using threshold-based techniques to analyze monsoon rainfall, since many models have dry biases, and thus fail to attain absolute rainfall amounts adequate for defining monsoon. In the fractional accumulation approach all of the models are interpreted within the context of their own annual cycle irrespective of absolute biases in amount. Figure 4a shows the All-India rainfall accumulations, indicating that there is large dispersion in the ability of the CMIP5 models to simulate the absolute rainfall amounts. The fractional accumulations, seen in Fig. 4b, reveal a systematic bias, with the monsoon onset being delayed in most models.

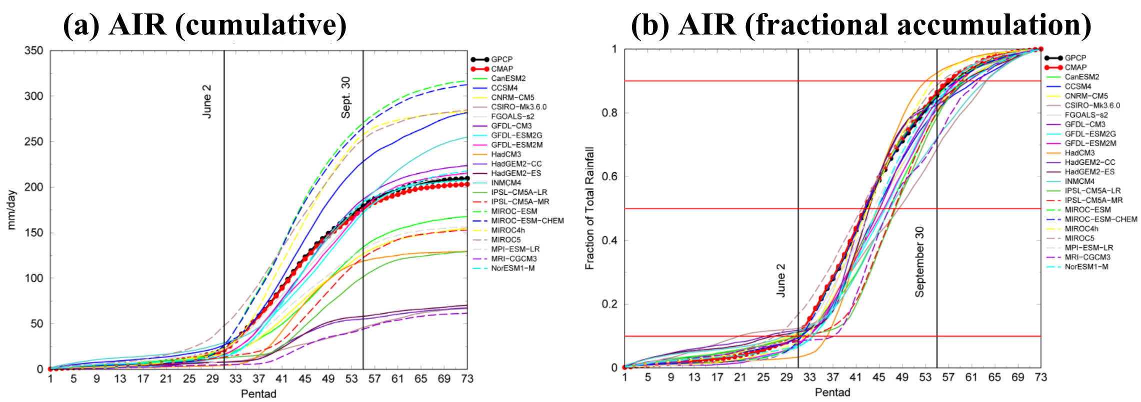


Figure 4: (a) Cumulative rainfall over India, and (b) fractional accumulated rainfall over India. (from Sperber and Annamalai 2014).

Over the Australian region (not shown, see Sperber and Annamalai 2014) nearly half of the models suggest an early onset while others suggest delayed onset. Additionally, models that have realistic fractional accumulation over India (e.g., MIROC-ESM) can fail to capture the annual cycle over Australia, and vice-versa (e.g., MRI-CGCM3). In summary, a complete north-south migration of the ITCZ that heralds the AAM system may be absent in many models or it may be too weak. Additionally, the rate at which precipitation is accumulated during the peak monsoon season can vary substantially across the suite of models (Sperber and Annamalai 2014). This suggests that the space-time characteristics of the rainfall are not well simulated. In summary, model systematic errors in simulating the annual cycle of the South Asian monsoon persist despite dedicated efforts.

3. Subseasonal variability of the South Asian summer monsoon

3.1 Intraseasonal (30-50 day) variability: The space-time evolution of the boreal summer intraseasonal variability (BSISV) is more complex than its boreal winter counterpart, the MJO. While the MJO is dominated by eastward propagation (Madden and Julian 1994), in the BSISV the equatorial component interacts with the mean monsoon flow resulting in north and northwestward propagating components over the northern Indian Ocean and tropical west Pacific, manifested as monsoon active-break phases (Lau and Chan 1986; Annamalai and Sperber 2005). As regards to BSISV, diagnostics on CMIP3 models suggest that one of the necessary conditions is that the models capture the equatorial component with sufficient amplitude for the generation of Rossby waves. The poleward migration of BSISV depends on the poleward extension of mean easterly shear and three-dimensional distribution of moisture (Wang and Xie 1997), and an idealized coupled model study suggests the role of anomalous horizontal moisture advection by the mean winds as one possible thermodynamic factor for the poleward migration (Ajayamohan et al. 2011). Modeling and theoretical studies suggest that “mean conditions” act as necessary if not sufficient conditions for the existence of BSISV.

Of the CMIP3 database, Sperber and Annamalai (2008) noted that only 2 out of 17 models represented BSISV but this was an improvement compared to the previous generation of models (Waliser et al. 2004). Compared to CMIP3, the multi-model-mean composite evolution of the BSISV in CMIP5 captures salient features (Sperber et al. 2013), an encouraging model improvement. Figure 5 shows the space-time evolution of BSISV as simulated by the MIROC-5 coupled model, and day 0 corresponds to maximum convective anomalies over the eastern equatorial Indian Ocean. Initial enhanced convective anomalies are noted along the equatorial Africa at day -15, extending into the western equatorial Indian Ocean by day -10, and further eastward extension amplification occur during day -5 and day 0. The combined Rossby-Kelvin pattern in enhanced convective anomalies is noticeable at day 5 with signatures of poleward migration over Indian longitudes and eastward extension into Maritime Continent with weak signatures of northwest propagation over tropical west Pacific during day 10 to day 20.

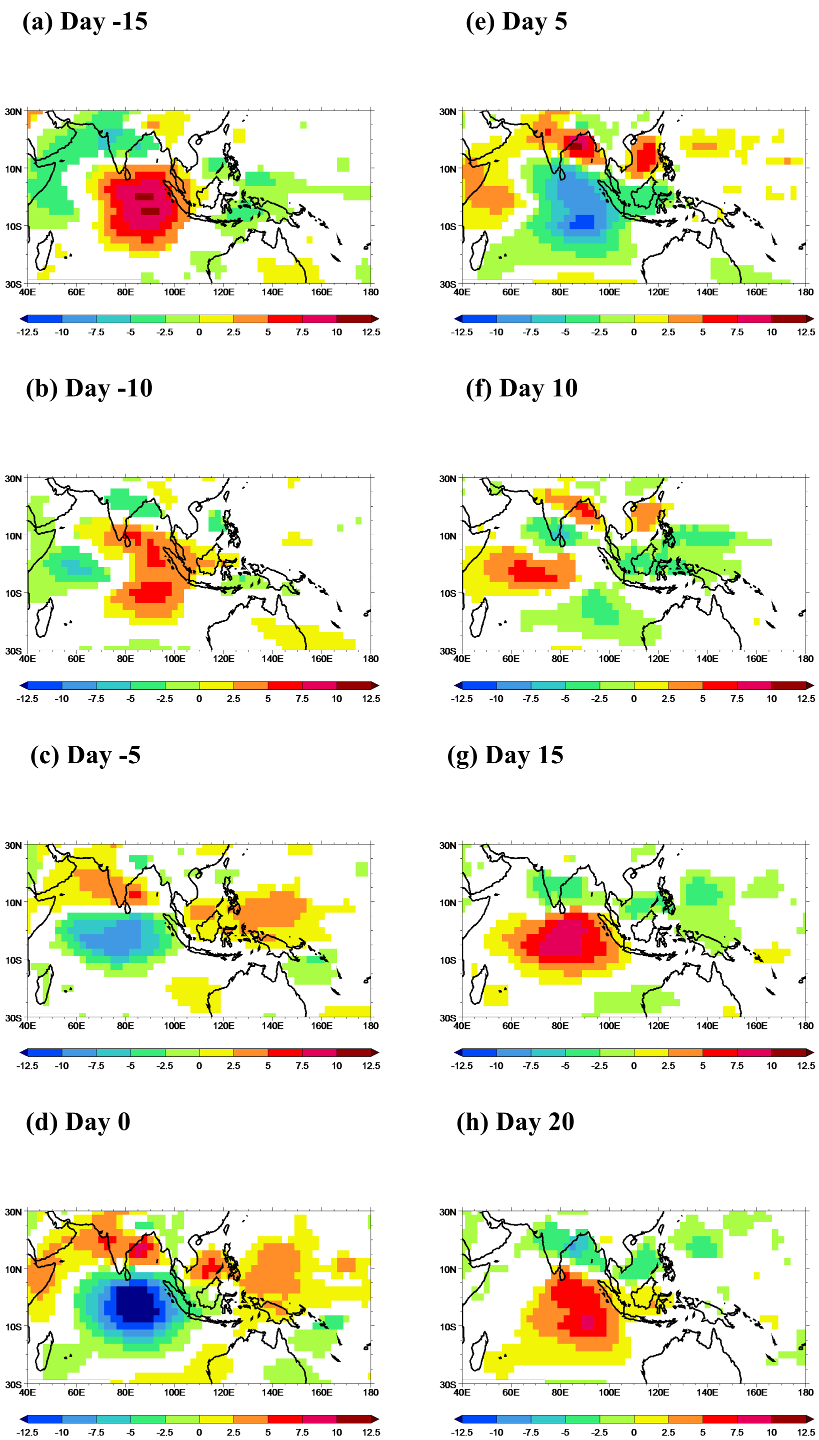


Figure 5: Lag regression of 20-100 day bandpass filtered OLR from MIROC5. The lag regressions have been scaled by one standard deviation to give units of W m^{-2}. Data are plotted where the regressions are statistically significant at the 5% level, assuming each pentad is independent.

At day -15 and day -10, one could note migration of suppressed convective anomalies over the Indian longitudes and into the Maritime Continent, and the cycle repeats. The space–time pattern clearly indicates that this mode is associated with northward propagation over the Indian longitudes and northwestward migration of convection over the tropical west Pacific. In Figure 5, the northward propagation occurs in conjunction with equatorial eastward propagation from the Indian Ocean to the west Pacific, characteristic of the MJO. One can easily delineate the coexistence of the three propagating components, and as such the intraseasonal modes are more complex during northern summer compared to northern winter. Yet, the fact that coupled models, even if only a very few of them, capture BSISV representation is encouraging and needs to be considered as a "boost" for model developers.

While active and break conditions are inherent to monsoon dynamics, prolonged dryness or extended breaks (breaks lasting 7 days or more) during the peak rainy season often result in droughts (e.g., Ramamurty 1969). Similarly, prolonged active monsoon conditions anchor local flooding. In observations, a break event is considered if, for three consecutive days rain- fall anomalies averaged over central India (21º–27ºN, 72º-85ºE) are below one standard deviation and persist for three consecutive days. Figure 6 shows the observed yearly statistics of monsoon breaks during boreal summer.

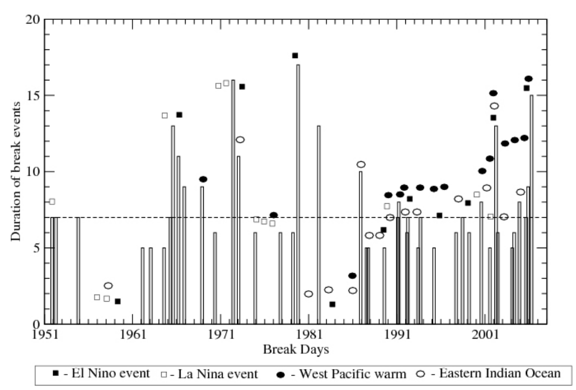


Figure 6: Yearly statistics of break days observed over central India. Years of anomalous sea surface temperature conditions over Nino3.4 region (filled squares for El Nino and open squares for La Nina), equatorial west Pacific (filled circles) and eastern equatorial Indian Ocean (90ºE-110ºE, 10ºS-0; open circles) are highlighted. The filled squares and open squares indicate JJAS Nino3.4 SST anomalies above and below one standard deviation, respectively. Similarly filled circles and open circles indicate JJAS SST anomalies above one standard deviation over the equatorial western Pacific and eastern equatorial Indian Ocean, respectively. The horizontal dotted line represents break days corresponding to 7 days (from Prasanna and Annamalai 2012).

While extended breaks occur in years of normal monsoon (e.g., 2000) and in neutral ENSO conditions (e.g., 1979), they tend to occur more frequently during years of warmer SST conditions over the equatorial Pacific and equatorial Indian Ocean. While the phase of BSISV leads to break conditions, the existence of "extended breaks" is anchored by seasonally persisting boundary forcing. Idealized numerical experiments performed with an AGCM supports the role of boundary forcing (Prasanna and Annamalai 2012) in which the frequency of occurrences of extended monsoon breaks over South Asia is significantly enhanced in response to El Nino forcing. Do climate models capture extended breaks?

To examine if a coupled model that displays realistic simulation of mean monsoon precipitation and aspects of subseasonal variability also capture extended breaks, we diagnosed all of the five-member ensemble simulations of the Geophysical Fluid Dynamics Laboratory (GFDL) Climate Model version 2.1 (CM2.1) coupled model. The criterion of a break event is similar to observations except that the averaging domain is considered over the whole of India (8º–28ºN, 63º–97ºE). For the models, the reason for choosing a different region is due to its limitation in simulating rainfall basic state and intraseasonal variability over central India (Sperber and Annamalai 2008). Figure 7 summarizes the occurrences of breaks lasting for three to seven days and more. Both in observations and model simulations, occurrences of short breaks (breaks lasting for three days) and extended breaks (breaks lasting for seven days or more) are higher, and it is encouraging that each model ensemble member captures this distribution.

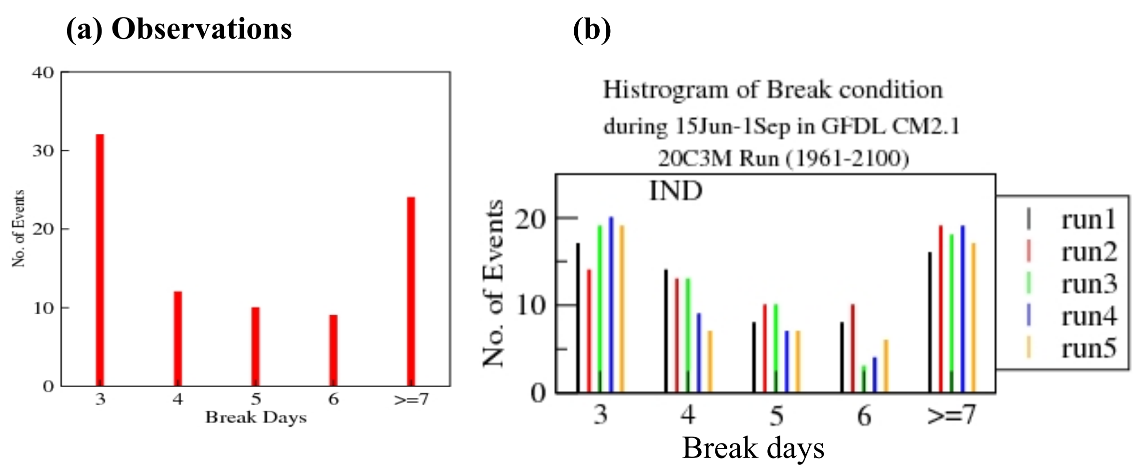


Figure 7: (a) Histogram of break days estimated from observed rainfall averaged over central India (21º–27ºN, 72º–85ºE); (b) as in (a) but from five-member CM2.1 integrations (1961–2000) and rainfall averaged over India (8º–28ºN, 63º–97ºE).

In summary, understanding and modeling the daily rainfall characteristics over central India involves: (i) understanding its statistical distribution (Fig. 6); (ii) identifying the processes from quality observations (dynamical and thermodynamical, and their interactions) and (iii) constraining model physics with field observations. It is, perhaps, not the lack of representation of a particular process but their partitioning that results from weaknesses in the various parameterization schemes that hinder skillful simulations of BSISV. A major stumbling block, in our opinion, is the lack of quality three-dimensional moisture and radiation observations over South Asia. Research community solely depends on reanalysis products to elucidate the processes, and in data sparse regions, model biases are severe in global reanalysis.

3.2 Synoptic systems: Boreal summer monsoon depressions, having a horizontal scale of about 2000-3000 km, form over the quasi-stationary monsoon trough. They are the most important components of the monsoon circulation. During the peak monsoon season (July- August), the majority of them forms over the warm waters of the northern Bay of Bengal, and move in a west-northwesterly direction with 4-6 systems forming each year (Sikka 2006). Over central India, the rainfall associated with the depressions contributes about 50% of the seasonal mean, and almost all extreme rainfall events there are associated with depressions (Sikka 2006). Therefore, to assess the future changes in the expected number of flood days, it is important to examine climate model projections of the depression strengths.

Observational and theoretical studies show that realistic mean conditions are a necessary, but not sufficient, condition for the formation and growth of monsoon depressions. The central-northern Bay of Bengal depressions develop in a region where the meridional gradient of potential vorticity (PV) vanishes (Shukla 1978). This PV behavior is one of the necessary conditions for the zonal jet instability mechanism of Charney and Stern (1962). Additionally, observational studies highlight the role of scale interactions since the BSISV modulates the formation of monsoon synoptic systems (e.g., Krishnamurthy and Ajayamohan 2010). Therefore, both the mean monsoon and space-time evolution of BSISV appear necessary conditions for the genesis and development of monsoon synoptic systems.

Only a few studies have evaluated the representation of monsoon depressions in climate models, partly due to resolution constraints and/or due to model systematic errors. Over the ASM region, in both NCEP-NCAR and ERA-15 reanalyses, Annamalai et al. (1999) noted that relevant statistics such as the growth/decay rate and the genesis/lysis locations are in good agreement between the two reanalyses, although the intensity is higher in ERA-15, which was performed at a higher spatial resolution (T106).

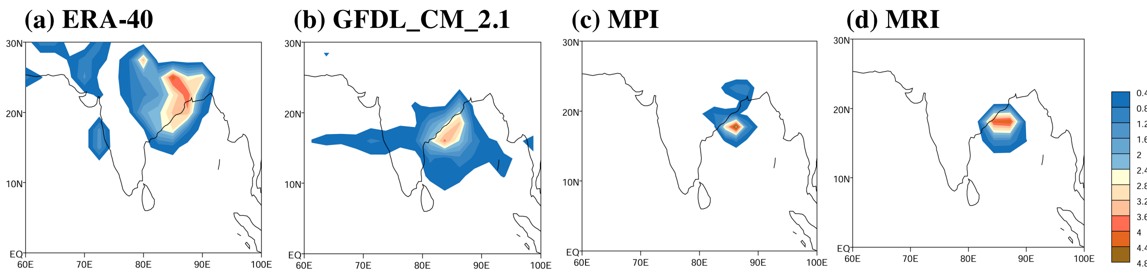


Figure 8: Occurrence of synoptic systems from: (a) ERA-40 data (1981-2000), the suite of twentieth century integrations of coupled models, (b) GFDL CM_2.1, (c) MPI, and (d) MRI. The last twenty years of the 20c3m simulations (1981-2000) are used for the calculation. The units are numbers per 2.5^{o} square box per 4-months (JJAS) period (from Stowasser et al. 2009).

In our earlier study, we examined the ability of select CMIP3 models to represent monsoon depressions (Stowasser et al. 2009), with the estimated number density for the period (1981-2000) summarized in Figure 8. The following criteria must be satisfied for a system that we identify and track: (i) a local vorticity maximum at 850 hPa exceeds 5.10^{-5} s^{-1} (Annamalai et al. 1999); (ii) a local pressure minimum exists within a radius of 250 km of the vorticity maximum; this minimum pressure is taken as defining the center of the storm system. To be considered as a model storm trajectory, a storm must last at least 2 days (see Stowasser et al. (2009) for details). The reanalysis confirms that most of the systems form north of 15^{o}N in the head Bay of Bengal and move inland in a westward to west- northwestward direction (Fig. 8a). Over the Arabian Sea, our analysis also captures the relatively less frequent mid-tropospheric cyclones whose signatures are seen at lower troposphere levels.

Although all three models represent the local maximum density over Bay of Bengal, the west-northwestward movement into the land regions is best captured by GFDL_CM2.1 (Fig. 8b). Compared to reanalysis, the genesis locations in the

GFDL_CM2.1 are shifted southwestward, and the model fails to capture the systems along the southern slope of the Himalayas. Despite its reasonable success in simulating the statistics of synoptic systems in the current climate, the GFDL_CM2.1 did not reveal any changes in the characteristics of monsoon depressions in a warmer climate (Stowasser et al. 2009). Bengtsson et al. (2006) found that the simulated depressions' intensity in ECHAM5 model is stronger than those in ERA-40 reanalysis while the track is closer to observations. Extending their analysis with ECHAM5 coupled model simulations the authors note an increase in the depressions' intensity in the 21st century compared to 20th century integrations. However, our analysis suggests that in the ECHAM5 coupled model studied by Bengsston et al. (2006) systems making landfall were not represented (Fig. 8c).

A striking result from observations is a steady decline in the formation of monsoon depressions since about 1970s (Sikka 2006). Additionally, there is no change in the formation of total number of synoptic lows over South Asia but their intensification into depression strength has rather declined. Our ongoing analysis with CMIP5 solutions is focused on identifying the possible processes for the declining tendency in monsoon depressions. Any future projections in the intensity of monsoon synoptic systems require that climate models capture not only their statistics but also the effect of any long term forcing.

4. Interannual variability

For the Asian monsoon, a successful prediction of the seasonal mean (June through September) rainfall anomalies helps in the planning of agriculture, hydroelectric and fresh water resources. With two-thirds of South and South East Asian work force being agro-related and a large portion of the cultivable land are rain-fed, seasonal forecasts of monsoon rainfall anomalies are increasingly scrutinized. Before models are employed for operational predictions, a question that needs to be systematically examined is: *Do the present-day coupled models faithfully represent the necessary ingredients to account for the interannual variations of monsoon, particularly severe strong and weak monsoons?*

Charney and Shukla (1981) hypothesized that the seasonal mean rainfall and circulation anomalies of the *large-scale* monsoon are governed by slowly varying boundary conditions, such as SST, snow cover, soil moisture etc., an hypothesis supported by many observational and modeling studies (e.g., Walker and Bliss 1932; Rasmusson and Carpenter 1983; Soman and Slingo 1997; Annamalai and Liu 2005; Lau and Nath 2000). For example, within available observations all severe weak monsoon years (15% below normal) over India co-occurred with El Niño (Pillai and Annamalai 2012). Of the boundary forcing elements ENSO dominates monsoon interannual variations, though the specific mechanism(s) through which ENSO influences monsoon remains an open question.

Sensitivity experiments (Turner et al. 2005) and diagnosis of a suite of models that participated in CMIP3 (Annamalai et al. 2007) suggest that realistic representation of the basic states in the tropical Pacific and rainfall over South Asia, as well as the timing and location of the diabatic heating anomalies associated with ENSO, are necessary conditions for simulation of the monsoon-ENSO teleconnection. t Additionally, at interannual time scales, the SST-rainfall relationship over the warm pool is complex (Wu and Kirtman 2005). Therefore, realistic simulation of regional SST variations is another *necessary condition*. Even if the above necessary conditions are met, examining the association in one realization (Annamalai et al. 2007) and/or a select period within one realization (Sperber et al. 2013) may not yield robust results since the variance of ENSO itself waxes and wanes at decadal-to-centennial time scales (Wittenberg 2009).

Annamalai et al. (2007) examined the ENSO-monsoon association in CMIP3 models. To check if the models represent the timing of the teleconnection correctly, lead/lag correlations between Niño-3.4 (5°S–5°N, 120°–170°W) SST anomalies and all-India rainfall (AIR) anomalies were computed. For each of the four models, this correlation is calculated separately for each realization, and then an ensemble-mean pattern is computed.

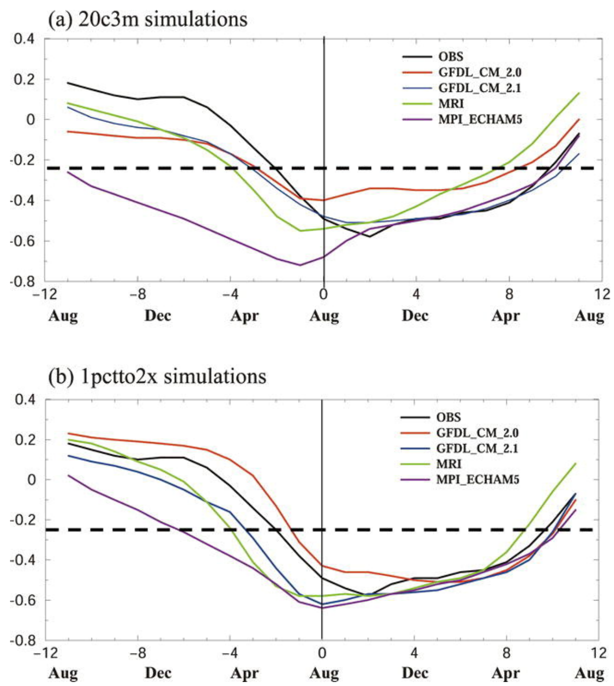


Figure 9: Lag–lead correlation between AIR anomalies and Niño-3.4 SST anomalies: (a) 20c3m and (b) 1pctto2x simulations. In (a) and (b), the results from observations are also shown. Horizontal dotted lines represent the 5% significance level. Lag–12 corresponds to Niño-3.4 SST anomalies one year before the monsoon season.

In observations negative correlations (Fig. 9a, black line) occur only after April. The observed maximum correlation after the monsoon season has led to suggestions that variations in the intensity of the monsoon can potentially influence the surface wind stress in the equatorial Pacific and thereby modify the statistical properties of ENSO (e.g., Kirtman and Shukla 2000). In the 20th century (20c3m) simulations, all of the models capture the inverse relationship during boreal summer, but the maximum negative correlation occurs too early in the GFDL_CM_2.0, MRI, and MPI_ECHAM5 simulations. While three of the models reasonably represent the spring predictability barrier, seen as the near-zero correlations during the preceding winter/spring, the

MPI_ECHAM5 simulations (violet line, Fig. 9a) are incorrect in this respect, with the presence of pronounced negative correlations from the preceding winter. Of the four models, GFDL_CM_2.1 best captures the timing of the relationship correctly. The ability to resolve the timing is possibly related to the ability in simulating the space–time evolution of SST and the associated diabatic heating anomalies during El Niño events. In the 1pctto2x integrations (Fig. 9b) the tendency is for the spring predictability barrier to be more apparent, and in the case of GFDL_CM_2.0 and MRI there is a tendency for the negative correlations to persist for about 3–6 months after the monsoon season. Overall, the results presented so far indicate that the ENSO–monsoon relationship remains strong and stable in a warmer climate.

In these four CMIP3 models, the future projection of south Asian monsoon interannual variability was studied (Fig. 10; Turner and Annamalai 2012) that depicted realistic mean monsoon precipitation and ENSO characteristics. The PDF is based on the pre-industrial control run (solid) and in the simulation where 1% /year increase in the concentration of CO_2 was imposed (dashed). The standard deviation in the control simulations is also given for each of the models.

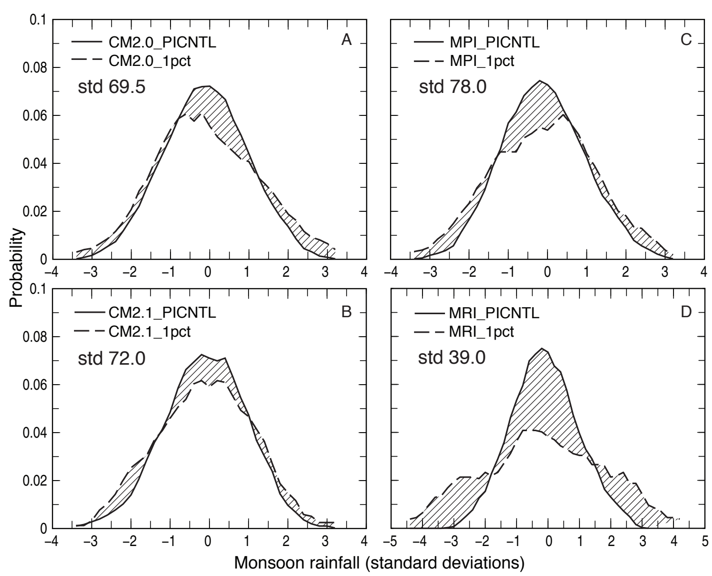


Figure 10: Probability density functions of interannual variability in monsoon rainfall in control and future climate scenarios. Shown are normalized probabilities of occurrences in four CMIP3 models. Preindustrial control (solid) and 1% per year increase in CO_2 concentrations (dashed) are shown. The future variations are scaled by preindustrial control interannual standard deviations. The differences in the shape of the PDFs have been tested for significance based on K-S test (after Turner and Annamalai 2012).

While all the four models suggest a reduction in the occurrences of normal monsoon years over South Asia, a statistically significant shift in the tail of the PDF is noted only for the MRI model (Fig. 10d), i.e., the response is very large in the model that has low monsoon variability (low standard deviation) in the control simulation. In summary, while it is encouraging that ENSO-monsoon association remains intact in a warmer planet, the uncertainties in the future projection of ENSO itself, to a degree, causes uncertainty in the future behavior of severe weak or strong monsoons over South Asia.

5. Longterm variability

During the last six decades (1950-2010), seasonal mean rainfall observations over India suggest for a declining trend (e.g., Ramanathan et al. 2005). While the drying trend may be due to factors other than increased GHG concentration, such as aerosols (Ramanathan et al. 2005), Annamalai et al. (2013) showed that anthropogenic forcing through SST warming over the tropical western Pacific likely caused an east-west shift in monsoon rainfall, with a drying tendency over South Asia. Unlike in future projections due to anthropogenic forcing where rainfall increases over both South Asia and tropical western Pacific (e.g., Turner and Annamalai 2012), in the last 6 decades, atmospheric response to SST warming over the tropical Indian Ocean has not yet occurred (see Annamalai et al. 2013 for more details).

To ascertain the spatial extent of drying over South Asia, the linear trend was estimated in gridded observational rainfall and analyzed circulation fields. In land-based rainfall products (e.g., Fig. 11a), the declining pattern over the plains of central India and Indo-China is consistent with other studies (e.g., Ramanathan et al. 2005). The spatial extent of drying is captured in other rainfall observations [not shown here, but see Bollasina et al. (2011)] but the estimated intensity is higher in CRU (Fig. 11a). From the results presented here and elsewhere, one robust feature is the declining tendency in monsoon rainfall over central India but because of observational uncertainties and different algorithms employed in re-gridding station rainfall observations, the magnitude of this declining varies among the products.

In seeking attribution for this declining tendency, Annamalai et al. (2013) estimated trends in SST and circulation fields. During 1949–2000, it is clear from Fig. 11c that SST rose in both the tropical Indian Ocean and the western tropical Pacific (a region extending over 10ºS–30ºN, 60º–150ºE)—in the former region by 0.75ºC, and in the latter by 0.5ºC. Yet, in the Indian Ocean, despite the SST rise, SLP has increased over the western Indian Ocean, and climatological southwesterly monsoon winds have weakened (Fig. 11b).

In the tropical western Pacific (10º–30ºN, 130º–170ºE), in contrast, SLP has dropped and the cross-equatorial flow emanating from the Australian high, the low-level westerlies, and anomalous cyclonic circulation have strengthened (Fig. 11b). Direct rainfall observations are unavailable; nevertheless, the decrease in sea surface salinity from 1955 to 2003 (Delcroix et al. 2007), the increased atmospheric water vapor content from 1988 to 2006 (Santer et al. 2007), and the drop in SLP (Copsey et al. 2006) all imply this region has had more rainfall in recent decades. Thus, observations point to a change in the monsoon circulation that has resulted in less rainfall over South Asia and more over the tropical western Pacific.

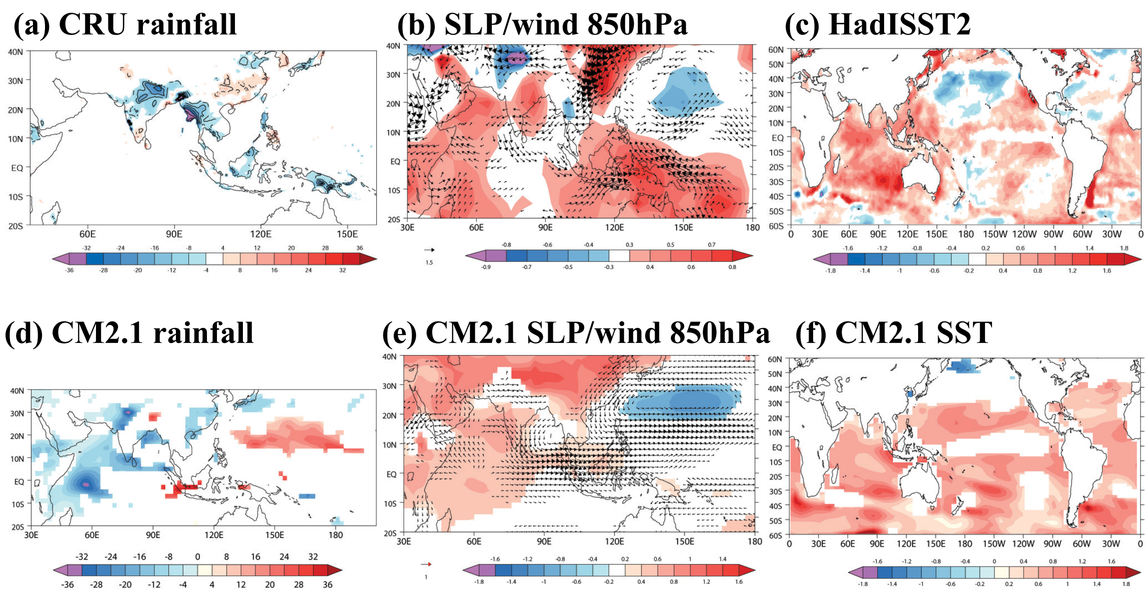


Figure 11: (a) Linear trend in observed rainfall [mm month^{-1} (52 yr)$^{-1}$] during boreal summer (JJAS) from the CRU gridded dataset for the period 1949–2000. (b) As in (a), but for sea level pressure [hPa (52 yr)$^{-1}$] from Hadley Centre data and 850-hPa wind [m s^{-1} (52 yr)$^{-1}$] from the NCEP–NCAR reanalysis. (c) As in (a), but for SST data from HadISST2 [8C (52 yr)$^{-1}$]. (d) as in (a) but from CM2.1. (e) as in (b) but from CM2.1 and (f) as in (c) but from CM2.1. In each of the panels, negative trend values are shown in blue and purple while positive values are in red. At each grid, only trend values greater than local interannual standard deviation are shown (adopted from Annamalai et al. 2013).

Is there support from a coupled model simulation for this observed east–west shift in monsoon rainfall? During 1949–2000, barring regional details, CM2.1 simulated the observed drying over South Asia (Fig. 11d), the increased SLP and the reduced monsoon circulation over the northern Indian Ocean (Fig. 11e), the increased SST in the Indo-Pacific warm pool, and the insignificant warming along the equatorial eastern Pacific (Fig. 11f). In the tropical western Pacific, increased rainfall (Fig. 11d) is dynamically consistent with stronger low-level cyclonic circulation and lower SLP (Fig. 11e). While the coupled model captures the broad features noted in observations and reanalysis (Figs. 11a-c), there are certain limitations including lack of simulation of positive rainfall trend over southern China and negative rainfall trend over the plains of Indo-China (Fig. 11d).

Turner and Annamalai (2012) examined the spatial pattern of the monsoon rainfall response to anthropogenic warming for the time-mean equilibrium response to increasing greenhouse-gas concentrations in only the 1pctto2x experiment for 20 CMIP3 models. The multimodel mean (see Fig. 3 in Turner and Annamalai 2012) results suggests enhanced rainfall over parts of South Asia and the same mean change computed for the four 'reasonable' models (models considered in Fig. 10) shows a similar result, providing more confidence in such a projection. The strengthened monsoon rainfall is generally attributed to increasing atmospheric moisture content over the warmer Indian Ocean, resulting in increased vertically integrated moisture fluxes towards India — such thermodynamic forcing has been consistently shown to lead to precipitation increases for South Asia.

In summary, climate models need to capture the "observed" east-west shifts in monsoon rainfall in the current climate for providing confidence in their future projections. A cursory examination with CMIP5 models (not shown) suggests that most models fail to capture the drying tendency over South Asia during 1950-2005 in their historical simulations.

6. Summary and future directions

It is now recognized that any future projections of long-term changes in climate can only by made by complex climate models that depend on super computers. The world modeling summit held in May 2008 argued that “climate models will, as in the past, play an important, and perhaps central role in guiding the trillion dollar decisions that the peoples, governments and industries of the world will be making to cope with the consequences of changing climate” (Shukla et al. 2009). Can the ultra-super computers alone solve monsoon-modeling issue?

It is fair to mention that in climate science there is no bigger problem than modeling the monsoons. Even after the availability of super computers, there are many other factors that will limit our ability to model the monsoons because the inherent complexity of the monsoon itself! For example, while the importance of large-scale forcing such as orography and boundary forcing such as SST are fairly understood (Hoskins and Rodwell 1995; Charney and Shukla 1981), the specific role of thermodynamic processes as well oceanic process in shaping the mean monsoon precipitation and its spectrum of variability are just beginning to unfold.

Furthermore, observational, theoretical, and modelling studies confirm that the mean monsoon precipitation and circulation influence monsoon variability on all timescales. Moreover, there is a suggestion that scale interactions with ENSO impact the frequency of occurrence of drought and flood monsoon conditions (>15% of the climatological normal; Pillai and Annamalai 2012),withwith extended active (break) monsoon conditions giving rise to flood (drought) years (Prasanna and Annamalai 2012). During flood years the frequency of occurrence of heavy rainfall events associated with synoptic systems is higher (Stowasser et al. 2009). Almost all heavy rainfall events (> 50 mm/day) are associated with synoptic systems that are largely punctuated by the phase of intraseasonal variability (Krishnamurthy and Ajayamohan 2010). Thus, the statistics of monsoon extremes are, *to a certain degree*, modulated by the phases of natural modes of variability. In summary, the processes involved in the “scale interactions” present in the monsoon system have not yet been understood, nor has their realistic representation in climate models been assessed.

We posed a few questions in the Introduction – all suggesting limitations in monsoon modeling. It is our contention that despite many books and numerous scientific publications on monsoons our knowledge in the monsoon behavior is still limited. A classic example is the ongoing monsoon season of 2014 in which break-like conditions have prevailed for over 5 consecutive weeks, and the June 2014 monthly rainfall anomaly is about 44% below normal over continental India (Source: India Meteorological Department. http://www.imdpune.gov.in/mons_monitor/mm_index.html) – and we do not yet know the reasons for this peculiar behavior. While the ongoing moderate strength developing El Nino in the equatorial Pacific is expected to exert an impact on the South Asian monsoon, the prolonged persistence of dry spell in the month of June is rather unique.

As alluded to in the Introduction, it is our view that high quality three dimensional observations of atmosphere (moisture, radiation, etc.) and ocean (temperature, salinity, etc.) over the monsoon influenced regions are essential for better understanding, modeling, and predicting the monsoon behavior. *Our conclusion is that without such an observational effort improving the physical processes in numerical models is severely limited.* As a consequence, this poses a severe constraint in improving skill in monsoon prediction with weather and climate models.

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